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# **Synchronized proxy-based temperature reconstructions reveal mid- to late Holocene climate oscillations in High Arctic Svalbard**

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*Running head:* Mid- to late Holocene climate oscillations in High Arctic Svalbard

## 26   **Abstract**

27

28   Existing paleoclimate data are exceedingly sparse from southern Spitsbergen, a High Arctic region  
29   predicted to experience significant environmental changes as a result of amplified warming. We  
30   analyzed biotic and isotopic paleolimnological proxies to reconstruct past climate from a lacustrine  
31   sediment core, with a basal age of ~5500 years BP, in southern Spitsbergen (77°N). We used fossil  
32   Chironomidae assemblages to quantitatively reconstruct past mean July air temperatures and stable  
33   oxygen isotope values ( $\delta^{18}\text{O}$ ) of these fossils to estimate changes in mean annual air temperature  
34   (MAT). These proxy records are strikingly similar and show that the coldest anomaly since the mid-  
35   Holocene occurred between 350-50 cal a BP, during the “Little Ice Age”, whereas the warmest  
36   period in the summer temperature record occurred between 5500-5000 and ~2000 cal a BP. Our  
37   findings indicate that the natural long-term air temperature dynamics in our study area are most  
38   likely connected to solar minima and positive feedback mechanisms from sea-surface temperature  
39   maxima. The results also highlight that the recent temperature increase is unprecedented in its rate  
40   with a ~2 °C increase in the summer temperatures during the past ~50 years.

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50   *Keywords:* Arctic; Chironomidae; Paleoclimate; Spitsbergen; Stable isotopes

## 51    **Introduction**

52

53    Arctic regions, such as Svalbard, are highly impacted by present day climate change due to multiple  
54    feedback mechanisms related to oceanic, cryospheric and atmospheric processes (Pithan and  
55    Mauritsen, 2014). Arctic sea ice extent and volume have been declining (Stroeve *et al.*, 2008;  
56    Polyak *et al.*, 2010), and the Arctic Ocean may become seasonally ice-free by the mid-21<sup>st</sup> century  
57    (Wang and Overland, 2009). It was recently demonstrated that the Atlantic Meridional Overturning  
58    Circulation (AMOC), which plays a key role in the poleward transport of heat and has a profound  
59    impact on climate in northwestern Europe and the High Arctic, may have slowed due to global  
60    warming (Landerer *et al.*, 2015). To improve our understanding of the Arctic climate change  
61    processes, a long-term perspective of the magnitude and rate of climate fluctuations in the past is  
62    necessary.

63            Arctic lakes and their sedimentary archives provide an opportunity to examine past  
64    environmental conditions. Paleolimnological methods, using biotic proxies such as chironomids  
65    (Insecta: Diptera), which are sensitive to temperature changes, and stable oxygen isotope signatures  
66    ( $\delta^{18}\text{O}$  values) reflecting hydroclimatic conditions, have been found to be very useful in tracking past  
67    climate changes (Brooks, 2006; Wooller *et al.*, 2004). Fossil chironomid assemblages have been  
68    used to reconstruct temperature during summer, which is their primary season for growth and  
69    development (Eggermont and Heiri, 2012), whereas the isotope values of their head capsules can  
70    capture the annual  $\delta^{18}\text{O}$  signal of lake water derived from precipitation (Verbruggen *et al.*, 2011).  
71    The  $\delta^{18}\text{O}$  of head capsules of chironomid larvae are equilibrated with the  $\delta^{18}\text{O}$  of lakewaters in  
72    which they live. In lakes that are not substantially influenced by evaporation or glacial meltwater  
73    inputs, the  $\delta^{18}\text{O}$  values of lakewater are controlled by the  $\delta^{18}\text{O}$  value of local precipitation and  
74    temperature. In these lakes chironomid  $\delta^{18}\text{O}$  values can subsequently be used to infer mean annual  
75    air temperature (MAT) (Wooller *et al.*, 2004).

76 In Svalbard, quantitative paleoclimatic records are thus far sparse, especially for the  
77 southern part of Svalbard, and cover a time span of  $\leq 2000$  years (Brooks and Birks, 2004; Isaksson  
78 *et al.*, 2005; D'Andrea *et al.*, 2012). There is a pressing need for long-term paleoclimate records  
79 from the area for linking climate and AMOC dynamics. In our study area, the North Atlantic  
80 Current carries warm surface waters from the Atlantic to high latitudes along the western side of  
81 Svalbard. Cold waters from the Arctic Ocean move southwards on the eastern side of Svalbard  
82 crossing the warmer waters at the southern tip of Svalbard, where the Polar Front develops  
83 (Majewski *et al.*, 2009). Our study area has a significant marine influence from the warm West  
84 Spitsbergen Current and the cold Sorokap Current, and is therefore a climatically and  
85 oceanographically sensitive region. For this reason, we examined a lake sediment profile from  
86 southern Spitsbergen (Fig. 1) using analyses of chironomid community assemblages to reconstruct  
87 mean July air temperature combined with  $\delta^{18}\text{O}$  values from analyses of chironomid head capsules to  
88 infer past MAT. The results of our study are important for understanding the long-term climate  
89 processes and land-ocean-atmosphere interactions in this High Arctic climate change hotspot area.

## 91 **Methods**

### 92 *Study site and sampling*

93  
94 Lake Svartvatnet (76°90'N, 15°66'E, 63 m a.s.l.) is located in a periglacial landscape (Arctic  
95 tundra) in southern Spitsbergen, across the Hornsund fjord from the Polish Polar Station (Fig. 1).  
96 This thermally stratified lake has a maximum depth of 26.5 m, surface area of ~80 ha and one outlet  
97 on its eastern side. Svartvatnet is an oligotrophic lake with a surface water pH of 7.2 (measured  
98 June 2013). The lake is annually ice-covered for 9-10 months with an open-season from mid-July to  
99 September. At present, the mean July temperature in the area is 4.5 °C, while the MAT is -2.1 °C  
100 with annual precipitation remaining below 400 mm (Marsz and Styszyńska, 2013). The lake

101 receives most of its water during the spring/summer as meltwater via a set a shallow creeks,  
102 whereas the periglacial soil appears not to produce significant groundwater storage at the study site  
103 (Ojala *et al.*, 2016).

104 A 167-cm sediment core (SV4C) and a parallel 149-cm core (SV4B) were taken from the  
105 northern basin of Svartvatnet (Fig. 1) using a light piston corer during the summer of 2013. The  
106 sediments consisted of clay gyttja with visible sand and silt/clay laminae (Ojala *et al.*, 2016). The  
107 sediment sequences were measured for their physical properties, correlated and the chronology was  
108 determined based on a combination of  $^{137}\text{Cs}$ ,  $^{210}\text{Pb}$ , AMS  $^{14}\text{C}$  (five dates) and paleomagnetic dating.  
109 The dating results and the constructed age-depth model (a Bayesian P-sequence deposition model)  
110 (Fig. S1) demonstrated that the SV4C sediment profile covers a 5500-year sedimentary archive with  
111 an average accumulation rate of  $0.3 \text{ mm a}^{-1}$  (Ojala *et al.*, 2016). The accumulation remained  
112 otherwise stable and continuous throughout the core but according to the  $^{210}\text{Pb}$  dating, the  
113 sedimentation rate increased during the 20<sup>th</sup> century. The well-established sediment chronology,  
114 with all the dating results showing good agreement, provides a solid basis for paleoclimatic  
115 reconstructions from the Svartvatnet sediment sequence. Detailed descriptions of the lake basin and  
116 the sediment, and the associated age-depth model for the Svartvatnet succession was previously  
117 published (Ojala *et al.*, 2016).

118

## 119 *Analyses*

120

121 Sediment subsamples (SV4C) for fossil chironomid analysis were prepared using standard methods  
122 (Brooks *et al.*, 2007). The wet sediment was sieved through a 100- $\mu\text{m}$  mesh and the residue was  
123 examined under a stereomicroscope. Larval head capsules were extracted with fine forceps and  
124 mounted permanently on microscope slides. Identification was performed under a light microscope  
125 at 400x magnification. The minimum chironomid head capsule number per sample was set to 50

126 (Heiri and Lotter, 2001). Principal component analysis (PCA) was used to detect primary direction  
127 of the chironomid community variance (Abdi and Williams, 2010). The PCA was run with all taxa  
128 included using a square-root transformation of the species percentage data. The chironomid-based  
129 mean July air temperature reconstruction (CiT) used the trimmed Norwegian weighted-averaging  
130 partial least squares (WA-PLS, 3-component) calibration model (Velle *et al.*, 2011). The model has  
131 a cross-validated coefficient of determination of 0.86, a root mean squared error of prediction of  
132 0.70 °C and a maximum bias of 0.66 °C. To test whether the reconstruction corresponded to the  
133 primary chironomid community variability, the temperatures were compared against the PCA axis 1  
134 scores using Pearson product-moment correlation coefficient and the associated level of statistical  
135 significance.

136         The protocol for  $\delta^{18}\text{O}$  analysis of chironomid head capsules ( $\delta^{18}\text{O}_{\text{chir}}$ ) followed Wang *et al.*  
137 (2008) without the acid treatment, and was described further by Arppe *et al.* (2017) and Kurki  
138 (2016). Approximately 80  $\mu\text{g}$  of cleaned chironomid head capsules were weighed in silver capsules,  
139 and measurements of  $\delta^{18}\text{O}$  values were performed with a Finnigan ThermoQuest TC/EA coupled to  
140 a Delta<sup>Plus</sup> Advantage isotope ratio mass spectrometer at the Laboratory of Chronology, Finnish  
141 Museum of Natural History. The  $\delta^{18}\text{O}$  values are expressed relative to the oxygen isotope ratio of  
142 Vienna Standard Mean Ocean Water. The international reference materials IAEA-NO3 (25.6‰),  
143 IAEA-601 (23.3‰), ANU sucrose (36.4‰), baleen whale keratin BWBII (14.0‰), and an  
144 internally validated IAEA-CH3 cellulose (32.6‰) were used to normalize raw  $\delta^{18}\text{O}$  values. The  
145 reproducibility of similar biogenic reference materials indicates a mean analytical precision (1  $\sigma$ ) of  
146 0.5‰, with a range from 0.4 to 2.5‰ depending on signal (sample) size, which is consistent with  
147 the findings of Wang *et al.* (2008). The  $\delta^{18}\text{O}_{\text{chir}}$  values were used to estimate mean annual values of  
148  $\delta^{18}\text{O}_{\text{lake water}}$  and MAT following Wooller *et al.* (2004) and Verbruggen *et al.* (2011). The composite  
149 errors for the reconstructions, factoring in both measurement uncertainty and calibration error, were  
150 calculated according to Pryor *et al.* (2014). They range from 1.4 to 3.6 ‰ (mean 1.6‰) and 1.9 to

151 4.5 °C (mean = 2.1 °C) for the  $\delta^{18}\text{O}_{\text{lake water}}$  and MAT reconstructions, respectively. A more detailed  
152 description of the analytical and data processing methods was described by Arppe *et al.* (2017). It  
153 should be noted that the  $\delta^{18}\text{O}_{\text{chir}}$  reconstruction of mean annual temperature should be considered  
154 tentative owing to reasons discussed below.

155

## 156 **Results**

157

158 Fossil chironomid assemblages of the initial part of the sediment profile (160-90 cm, ~5500-3300  
159 cal a BP) were dominated by *Micropsectra contracta*-type (Fig. 2), which corresponds to  
160 morphologically and ecologically similar taxon *M. insignilobus*-type in the chironomid-based  
161 calibration set used for the temperature reconstruction (Brooks and Birks, 2000, 2001). This taxon  
162 complex is typical for the oligotrophic lakes in Svalbard (Luoto *et al.* 2011; Velle *et al.* 2011). In  
163 the mid-part of the profile (90-50 cm, ~3200-1800 cal a BP) the cold-indicating *Hydrobaenus*  
164 *lugubris*-type began to dominate but was later (at ~1800 cal a BP) replaced by another cold  
165 preferring taxon *Oliveria*. In the most recent sample representing the present, several warm-water  
166 chironomids, such as *Paratanytarsus austriacus*-type and *M. junci*-type, increased that has also  
167 been detected in other lakes in Svalbard (Velle *et al.* 2011; Kivilä 2014).

168 In general, the chironomid PCA axis 1 scores were negative in the lower part of the record  
169 but turned into positive scores after ~3500 cal a BP. Reflecting a major shift in chironomid  
170 assemblages (decreases in *M. contracta*-type and *Orthocladius trigonolabis*-type and increase in *H.*  
171 *lugubris*-type), chironomid PCA axis 1 scores showed a rapid change at ~3300-3200 cal a BP (~90  
172 cm) and another change of similar magnitude in the most recent samples (decreases in *Oliveridia*  
173 and *H. lugubris*-type and increase in *M. junci*-type) (Fig. 2). The chironomid PCA axis 1 scores  
174 correlated closely ( $r=0.91$ ,  $P<0.001$ ) with the CiT changes during the past 5500 years (Fig. 3).  
175 Furthermore, the CiT values were very similar to the  $\delta^{18}\text{O}_{\text{lake water}}$  values and the  $\delta^{18}\text{O}_{\text{chir}}$



176 reconstruction of annual temperatures (Fig. 3). The CiT remained at  $>5^{\circ}\text{C}$  between  $\sim 5500\text{-}3500$  cal  
177 a BP with a progressive decline, however, with a short warm period at  $\sim 1500$  cal a BP and very  
178 warm temperatures at the surface sample. The reconstruction of  $\delta^{18}\text{O}_{\text{lake water}}$  values distinctly  
179 mimics the CiT reconstruction but the warmer temperatures between  $\sim 5500\text{-}3500$  cal a BP  
180 identified by the CiT are not as well reflected in the isotopic reconstruction. The minimum  
181 temperatures were reconstructed for the period between  $\sim 300\text{-}200$  cal a BP with both proxies. In  
182 addition to this temperature minimum, simultaneous cold periods were found at  $\sim 3400\text{-}3200$  and  
183  $1400\text{-}1200$  cal a BP.

184         The surface sediment CiT of  $6.6^{\circ}\text{C}$  is overestimated compared to the observed long-term  
185 mean July air temperature (1978-2014) at the study area of  $4.5^{\circ}\text{C}$  but close to the temperature  
186 measured in June 2016 of  $6.0^{\circ}\text{C}$ . Therefore, when compared to the present instrumental  
187 temperature record, the surface sediment CiT is within the model's prediction error of  $0.7^{\circ}\text{C}$ . The  
188 observed long-term MAT in the area is  $-4.4^{\circ}\text{C}$  being clearly lower than in the  $\delta^{18}\text{O}$ -based  
189 reconstruction for the surface sample. However, the reconstructed surface sample MAT of  $-0.6^{\circ}\text{C}$   
190 is closer to the last five-year mean of  $-1.9^{\circ}\text{C}$ , which remains within the reconstruction's sample-  
191 specific error estimate of  $3.0^{\circ}\text{C}$ . When comparing the reconstructed  $\delta^{18}\text{O}_{\text{lake water}}$  of  $-9.2\text{‰}$  (sample-  
192 specific error  $2.4\text{‰}$ ) with the observed lake water value of  $-9.6\text{‰}$ , measured from water samples  
193 during the summer 2013, the values were found to be very close.

194

## 195 **Discussion**

196

### 197 *Temperature reconstructions and climate variability*

198

199 According to our previous assessment based on contemporary aquatic communities (Luoto *et al.*,  
200 2016), the most significant confounding factor for chironomid-based temperature reconstructions in

201 the study area (Hornsund) is the potential influence of bird-driven nutrient enrichment. This kind of  
202 arctic eutrophication superimposed on the recent climate warming is a typical trend for smaller  
203 ponds lying close to shoreline in Svalbard with increasing sizes of arctic-nesting birds in their  
204 catchment (Luoto *et al.*, 2014, 2015). However, when comparing the Svartvatnet chironomid  
205 stratigraphy (Fig. 2) with the modern dataset (Luoto *et al.*, 2016), the phenomenon of bird-induced  
206 eutrophication is not reflected in the chironomid communities of Svartvatnet. Therefore, since our  
207 study site is not significantly influenced by birds, it is probable that the long-term biotic record is  
208 driven by the direct climate impact. In fact, the PCA results indicated that the chironomid  
209 assemblages significantly responded to temperature changes during the past 5500 years, as the  
210 chironomid PCA axis 1 scores correlated with the CiT. Furthermore, the  $\delta^{18}\text{O}_{\text{lake water}}$  reconstructions  
211 showed similar trends with the CiT (Fig. 3) providing evidence that both proxies were reflecting the  
212 same environmental variable (temperature), though at different timescales (annual vs. July).

213 While we consider the  $\delta^{18}\text{O}_{\text{lake water}}$  reconstruction a robust representation of past changes in  
214  $\delta^{18}\text{O}$  values of lake Svartvatnet water and local precipitation, the interpretation of the isotope record  
215 in terms of temperature variability is not straightforward. Due to uncertainties in both the present-  
216 day and past  $\delta^{18}\text{O}$ /MAT relationships, possible past changes in the locations and pathways of major  
217 moisture sources, and the seasonal distribution of precipitation, the  $\delta^{18}\text{O}_{\text{chir}}$ -based MAT  
218 reconstruction is subject to some uncertainty. The presented MAT reconstruction (Fig. 3) holds true  
219 only in a scenario where all the changes observed in  $\delta^{18}\text{O}_{\text{lake water}}$  can be attributed to changes in  
220 MAT at the site, without associated changes in the abovementioned factors. Indeed, the record of  
221  $\delta^{18}\text{O}$  values, while carrying a strong temperature dependent signal, is very likely also influenced by  
222 other environmental factors such as variations in sea ice extent around Spitsbergen, and should not  
223 be considered solely to reflect MAT. A detailed discussion of the  $\delta^{18}\text{O}$  record was previously  
224 presented by Arppe *et al.* (2017).

225 The reconstructions indicated relatively warm climatic conditions at the beginning of the  
226 record, between ~5500 and 3800 cal a BP. This warm period corresponds with the presence of  
227 smaller-than-present glaciers in Svalbard (Humlum *et al.*, 2005) and Franz Josef Land (Lubinsky *et*  
228 *al.*, 1999), and with the general timing of the Holocene Climate Optimum (HCO, ~8000-5000 cal a  
229 BP) (Wanner *et al.*, 2008). Our results are also in accordance with recent sediment-based Holocene  
230 glacier reconstruction from the northwestern coast of the Svalbard Archipelago (Røthe *et al.*, 2015),  
231 which shows low inorganic detrital input into the proglacial Kløsa basin, suggesting that influx of  
232 meltwater ceased following the possible melting of the Karlbreen glacier. Likewise, Svendsen and  
233 Mangerud (1997) concluded that the Linnébreen glacier at the mouth of Isfjorden in Spitsbergen  
234 had melted during this period. Based on plant macrofossil analysis, July temperatures may have  
235 been up to 1.5 °C warmer compared to the late 20<sup>th</sup> century in the western coastline of Svalbard  
236 (Birks, 1991). Our CiT shows that the summer temperatures between 5500 and 5000 cal a BP in  
237 southern Spitsbergen were similar to present day, which is in agreement with most paleoclimate  
238 records from Svalbard indicating generally warm mid-Holocene summer conditions (Lubinsky *et*  
239 *al.*, 1999; Humlum *et al.*, 2005; Luoto *et al.*, 2011).

240 Our CiT record, reflecting summer conditions, shows a general Neoglacial cooling trend  
241 from ~5000 cal a BP onwards (Fig 3). Svendsen and Mangerud (1997) suggested that following the  
242 HCO, the Linnébreen glacier reoccupied the valley above Linnévatnet between 4600 cal a BP and  
243 around 2800 cal a BP, probably in response to a decline in summer insolation and climate cooling.  
244 The first one of these episodes is fairly coincidental with two minimum peaks in our CiT. On the  
245 other hand, the  $\delta^{18}\text{O}_{\text{chir}}$  record, which incorporates the climatic signal of the cool/cold season, does  
246 not show a decline. Summer cooling, from a slightly later peak (4250-4050 cal a BP) than in our  
247 record, was recorded by van der Bilt *et al.* (2015) in the Lake Hajeren proglacial sedimentary record  
248 (Fig. 3). This may reflect regional differences caused by external forcing mechanisms, or slower  
249 response times of small glaciers (van der Bilt *et al.*, 2015).

250 We document a significant cooling episode in both our proxy records at ~3300 cal a BP,  
251 which has previously been detected in the equilibrium-line altitude (ELA) reconstructions from  
252 Karlbreen glacier on the northwest coast of Svalbard (Røthe *et al.*, 2015). Although our evidence is  
253 mostly based on single data point excursions (Fig. 3), a contemporaneous centennial-scale glacier  
254 advance in the Hajaren catchment was also observed by van der Bilt *et al.* (2015), which was linked  
255 to orbitally forced summer cooling and increased sea-ice export through the Fram Strait. In  
256 addition, a cooling event in our record at 1300 cal a BP is concurrent with the ELA reconstruction  
257 (Røthe *et al.*, 2015) and the Linnévatnet record (Svendsen and Mangerud, 1997), suggesting that the  
258 glacier response is probably more related to summer air temperature than precipitation changes.  
259 Peak erosion due to the presence of a glacier in the Lake Hajaren catchment occurred later, at  
260 around 1100 cal a BP (van der Bilt *et al.*, 2015) suggesting that the lack of glacier growth prior to  
261 1100 cal a BP was due to the climate being too dry or too warm in this low-lying catchment.

262 Typical for lacustrine sites in the Arctic (Miller *et al.*, 2010), the present results do not show  
263 clear signs of the Medieval Climate Anomaly (MCA, ~850-1250 CE). The closest study of  
264 millennial-scale climate fluctuations to our site was presented by Pawłowska *et al.* (2015), who  
265 investigated mass balance and glacial activity in a marine environment at Hornsund fjord. These  
266 investigators were not able to identify clear changes in environmental conditions during the MCA  
267 as there was a stable and low rate of sediment accumulation and low ice rafted debris flux.

268 According to our proxies, the general late Holocene cooling trend culminated at ~300-200  
269 cal a BP, coinciding with the end of the Little Ice Age (LIA). The previous ~2000/1000-year  
270 paleoclimate reconstructions from Svalbard indicate that the LIA was a rather muted and short-lived  
271 event in the area (Velle *et al.*, 2011; D'Andrea *et al.* 2012) compared to continental Northern  
272 Europe, where it lasted for several centuries (Luoto and Nevalainen, 2015, 2016; Zawiska *et al.*,  
273 2017). This is also well reflected in glacier ELA records, since the LIA is well represented in a  
274 record from continental Norway (Bakke *et al.*, 2005), while less distinct in the Svalbard glacier

record (Røthe *et al.*, 2015), suggesting that short-lived climate events may not be evident in glacier growth. According to our results, the declining trend towards colder LIA conditions may have initiated as early as 700-800 cal a BP, followed by a major cooling shift around 400 cal a BP. This date is in good accordance with the composite of high-resolution proxy climate records from the Arctic by Kaufman *et al.* (2009). It also agrees with the inferred beginning of the LIA at around 1600 CE described by Pawłowska *et al.* (2015) based on the analysis of foraminiferal ancient DNA in the marine environment of Hornsund fjord. Timing of the onset of the LIA at around 700 cal a BP also agrees well with other records from western Svalbard (Svendsen and Mangerud, 1997; van der Bilt *et al.*, 2015) suggesting a regional climate shift.

Our CiT for the surface sample is 3.5 °C higher than the lowest LIA values, whereas the  $\delta^{18}\text{O}_{\text{chir}}$ -based reconstruction, if interpreted to solely represent changes in MAT at the site, is >10 °C higher. However, the dramatic negative shift observed in the reconstruction of  $\delta^{18}\text{O}$  values during the LIA is likely to also reflect the significantly expanded sea ice cover limiting the input of proximally derived moisture, and thus should not be interpreted purely on the basis of MAT. These results nonetheless emphasize the influence of winter months to the annual heat budget in our study area. The annual and summer temperature increase from the LIA to the present appears to have been unprecedentedly rapid over the last 5500 years and clearly reflects the most significant climate shift in the record. As in previous evidence from the circum-Arctic (Kaufman *et al.*, 2009; Luoto *et al.*, 2017), rapid recent warming reverses the general trend of cooling in our reconstructions (Fig. 3). Our CiT record indicates a temperature increase of ~2 °C for the recent decades that is equal to the measured temperature increase during the past 30 years at the Polish Polar Station since 1979 (Marsz and Styszyńska, 2013). The surface sediment CiT of 6.6 °C is higher than the observed long-term mean July air temperature (1978-2014) at the study area of 4.5 °C but close to the present day (2016) temperature of 6.0 °C confirming that the reconstructed values are realistic. Similarly, the  $\delta^{18}\text{O}_{\text{chir}}$  record indicates a MAT value of -0.6 °C for the surface sample that is higher than the

300 35-year mean of 4.4 °C but closer to the last five-year mean of -1.9 °C. The recently (2013)  
301 measured  $\delta^{18}\text{O}_{\text{lake water}}$  value (-9.6 ‰) and the inferred  $\delta^{18}\text{O}_{\text{lake water}}$  value (-9.2 ‰) are also very  
302 close. Hence, the better comparability of the reconstructed values with the more recent observations  
303 are most likely owing to the observed rapid temperature increase over the temperature monitoring  
304 period (Marsz and Styszyńska, 2013).

305

### 306 *External forcing and feedback mechanisms*

307

308 Decreasing summer solar insolation may have caused the Neoglacial cooling trend in the Arctic  
309 region (Wanner *et al.*, 2008). In the North Atlantic region, the Neoglacial cooling is reflected by  
310 growing glaciers (Bakke *et al.*, 2010), decreasing sea surface temperatures (SSTs) (Sarnthein *et al.*,  
311 2003; Berner *et al.*, 2010) and ice-core derived temperatures (Lecavalier *et al.*, 2013). In general,  
312 North Atlantic surface winds and surface ocean hydrography have been influenced by variations in  
313 solar output throughout the Holocene, hence, the surface hydrographic changes may have affected  
314 production of North Atlantic Deep Water, potentially providing an additional mechanism for  
315 amplifying the solar radiation fluctuations and transmitting them globally (Bond *et al.*, 2001).

316 The air temperature dynamics in our study appear to be externally forced by solar minima  
317 and positive feedback mechanisms related to SST maxima. The cold episodes reconstructed from  
318 the current biotic and isotopic records at 3400-3200 and 1400-1200 cal a BP are synchronized with  
319 periods of low solar irradiation (Steinhilber *et al.*, 2009; Fig. 3), however, it is likely that  
320 combination of several factors triggers the cold and warm episodes. In agreement with Bond *et al.*  
321 (2001), our record shows that the LIA is not fully synchronized with the Maunder solar minimum.  
322 Cooling during the LIA is significantly connected with atmospheric forcing since a strongly  
323 negative NAO index phase prevailed (Trouet *et al.*, 2009; Dylmer *et al.*, 2013; Luoto and  
324 Nevalainen, 2017). The warm periods in our summer temperature record, ~5400, ~2200 and ~1600

cal a BP, show correspondence with increases in summer SSTs reconstructed from the northern Norwegian Sea (Sarnthein *et al.*, 2003; Fig. 2). However, these warm phases are not apparent in the isotopic record. Though this might reflect the sensitivity of  $\delta^{18}\text{O}$  values to precipitation (Langebroek *et al.*, 2011), it more likely mirrors its sensitivity to MAT. The lack of the warm MCA signal in our record is probably due to dominance of Arctic surface water in the western Barents Sea at this time, which contributes to cooler conditions than in continental regions of the Northern Hemisphere (Dylmer *et al.*, 2013).

The recent and most significant temperature increase in our record corresponds to the human-induced increase in greenhouse gases in the atmosphere since the pre-industrial age (Crowley, 2000). Although there is an increase in solar irradiance towards the present (Steinhilber *et al.*, 2009; Fig. 3), it appears to be too small to explain the fast and major increase in the temperatures at our study site. There also exists a possibility of a regional temperature response via ocean-atmosphere heat transport that has been suggested from marine sediments off Western Svalbard (Spielhagen *et al.*, 2011).

## Conclusions

Changes in fossil chironomid assemblages and  $\delta^{18}\text{O}$  values of chironomid head capsules document significant paleoclimate changes during the past ~5500 years in Svalbard. Both proxies showed similar temperature trends, indicating late Holocene cooling and the LIA. The recent temperature increase is unprecedented in its speed and magnitude since the mid Holocene. Although partly associated with single data point excursions, we document connections between periods of cold temperatures and solar radiation minima and warm periods and high SSTs, suggesting that both forcing/feedback mechanisms may play an important role behind the climate dynamics of our study region. The complexity of the regional climate patterns is reinforced by the local glacier influence.

350 The current record makes a valuable contribution to paleoclimate science, since no previous  
351 quantitative estimates of Holocene temperature changes are available from southern Spitsbergen,  
352 which is strategically located at the crossing of major warm and cold oceanic currents and where the  
353 Polar Front develops. Therefore, knowledge of climate trends and the processes controlling them at  
354 our study site provides significant understanding for circum-Arctic and global climate dynamics.

355

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357

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363

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572 **Figure captions**

573

574 **Figure 1.** Map. Lake Svartvatnet (76.895°N, 15.676°E, 63 m a.s.l.) is located 12 km south of the  
575 Polish Polar Station on Hornsund fjord, Svalbard.

576

577 **Figure 2.** Chironomid assemblages. Relative abundance of chironomid taxa and the principal  
578 component analysis (PCA) axis 1 scores from the Svartvatnet sediment sequence.

579

580 **Figure 3.** Reconstructions. Principal component analysis (PCA) axis 1 scores for fossil chironomid  
581 assemblages, chironomid-based mean July air temperature reconstruction (CiT), lake water  $\delta^{18}\text{O}$   
582 values derived from fossil chironomid head capsules and annual air temperature reconstruction  
583 based on the isotope records in Svartvatnet. The  $\delta^{18}\text{O}$ -based reconstruction, represented as  
584 anomalies from the record mean, should not be considered directly as a MAT record (discussed by  
585 Arppe *et al.* (2017) in more detail). The Svartvatnet data are compared with proxy records of  
586 average total solar irradiance (TSI) relative to the solar cycle minimum of the year AD 1986  
587 (Steinhilber *et al.*, 2009), summer sea-surface temperature from northern Norwegian Sea (Sarnthein  
588 *et al.*, 2003) and glacier variability in Svalbard (van der Bilt *et al.*, 2015). Cold climate episodes are  
589 marked in gray.

590

591 **Supporting Information Figure S1.** Age-depth model originally published in Ojala *et al.* (2016).  
592 Dating of the topmost sediments is based on  $^{210}\text{Pb}$  and  $^{137}\text{Cs}$  (solid black line). Calibrated AMS  $^{14}\text{C}$   
593 dates with 95.4% confidence levels (2s) are given in darker gray, black crosses represent their 2s  
594 median probability and black dots represent paleosecular variation dates. The age-depth relationship  
595 shaded with light gray is based on a Bayesian P-sequence deposition model.